Evidence from Caustic Waveform Modeling for Long Slab Thickening above the 660-km Discontinuity under Northeast Asia: Dynamic Implications

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ABSTRACT

Knowledge of the thickness and age of slabs is of great importance to further our understanding of the deformation of the subducting lithosphere and, in general, of regional mantle rheology. The technique of seismic tomography has been widely used to identify thermally induced mantle heterogeneities. However, because of the contamination of complicated caustic waveforms, its resolving power is relatively poor when it comes to image slabs in the mantle transition zone. In this study, we aim at constraining the thickness of the slab in the mantle transition zone beneath northeast Asia through the method of caustic waveform modeling for both P and S waves. We detect a high-velocity layer associated with a thickened slab lying in the transition zone and estimate its thickness to be ~140±20 km, corresponding to a thickening of 50–60 km at the long-lying segment. Numerical simulations indicate that downgoing material may be piling up under northeast Asia. The part of the Pacific plate subducting at the Japan trench thus can be interpreted to be in a buckling/folding retreating mode. We argue that thickening of the slab due to buckling instabilities could partly explain the enigma of the missing Pacific slab from the Cenozoic era. In addition, the slab might be close to the threshold of instability. Based on numerical simulations, mantle circulation may become temporarily layered. Cold and dense lithospheric material, after accumulating at the 660-km discontinuity, can suddenly sink into the lower mantle, resulting in an avalanche that, perhaps, might occur in the next tens of millions of years. This will be accompanied by a resurgence of volcanic activity, precipitated by upwellings emanating from the lower mantle close to the site of the avalanche.

1.1. INTRODUCTION

Subducted oceanic slabs represent the major source of buoyancy driving mantle flow [e.g., Ricard et al., 1993]. Characterizing the process of subduction is crucial to better understand the thermal and chemical evolution of our planet. In the past decades, seismic-wave tomography, based mainly on travel-time analysis, has proved able to resolve subduction-related thermal anomalies [e.g., van der Hilst, 1991; Fukao et al., 2001; Huang and Zhao, 2006; Li et al., 2008; Fukao and Obayashi, 2013]. However, intrinsic problems of damping-associated inversion techniques used in tomography as well as the poor resolution
at depth inhibit the possibility to place robust constraints of the detailed structure of subducting slabs. A recent study devoted to quantifying the uncertainty in travel-time tomography reveals that the root mean square velocity perturbations for many acceptable models compatible with the same travel-time dataset can vary from 0.3% to 1.3% in the upper mantle [de Wit et al., 2012]. Furthermore, in the transition zone, the limited ray coverage of the first arrivals caused by waveform triplication results in an even more ambiguous image.

Constraining the thickness of subducted slabs is of great importance in understanding problems related to mantle rheology and evolution of the lithosphere-mantle system. Slabs can undergo large deformation: they can bend, stretch, and thicken [Gurnis and Hager, 1988; King, 2001]. In the upper- to midlower mantle, thick blobs of seismically fast anomalies have been identified by different authors in subduction regions beneath Tonga, Marianas, and Kuril [e.g., Creager and Jordan, 1986; van der Hilst, 1995; Fukao and Obayashi, 2013].

Scaling laws derived from the theory of buckling of viscous sheets have been successfully employed to explain the apparent thickening of the subducted lithosphere beneath the transition zone as revealed by a few seismic studies [Ribe et al., 2007]. Fully numerical simulations of viscous flow with composite rheology have been performed to further explore the possible mechanisms leading to the emergence of long-wavelength thickening of slabs in the lower mantle [Béhouneková and Čížková, 2008]. Furthermore, 3D numerical models have shown that a relatively weak slab, with a viscosity not more than 100 times larger than that of the ambient mantle, can achieve a broad variety of shapes [Loiselet et al., 2010]. In spite of this focus on slab behavior in the middle to lower mantle, different styles of slab deformation have been identified in the upper mantle transition zone. The thickness of the subducted lithosphere in the upper part of the Bullen region [Bullen, 1963], which was proposed to be the transition region between the upper and lower mantle, however, is not well resolved. A slab with an initial thickness of \(~80–90\) km [Zhao et al., 1994] has been imaged beneath the Japan trench, behind which a vast amount of material lies subhorizontally over the 660-km discontinuity [e.g., Huang and Zhao, 2006; Li and van der Hilst, 2010, Fukao and Obayashi, 2013] (hereafter referred to as the 660). The resolved high-velocity layer in the stagnant slab, however, seems to occupy the whole mantle transition zone with an uncertainty of around \(\pm 90\) km or even more.

In this paper, we discuss our finding of a thickened slab in the upper mantle transition zone beneath northeast Asia and the possible mechanism of its formation, and we speculate on the dynamic implication associated with its fate. The constraint on the thickness of the slab is based on P and S wave models of Li et al. [2013] obtained from the method of caustic waveform modeling. We first show the sensitivity of the method to the structure around the 660-km discontinuity, and then briefly introduce the data sources and main features of the resolved P and S velocity models. New waveform datasets are added to verify the correctness of the obtained model, and a series of uncertainty tests based on waveform matching are performed, which provide a robust constraint on the thickness of the stagnant segment of the slab. 2D finite-volume numerical simulations were also performed in order to understand the dynamic implications of a thickened stagnant slab in the transition zone. We will argue that thickening of the slab due to buckling instability may partly explain the enigma of the missing Pacific slab. The slab may be close to be unstable, and the continuous piling up of lithospheric material may trigger an event of mantle avalanche beneath northeast Asia.

1.2. CAUSTIC WAVEFORM MODELING AND DATA SOURCES

Caustic waveforms are caused by seismic triplication [Aki and Richards, 2009], occurring in the presence of a first-order discontinuity with positive velocity jump or a sharp increase in velocity gradient (Figure 1.1). Because the pathways of triplicate phases lie very close to each other in the lithosphere and crust, differences in relative time and amplitude between different seismic branches, and especially the location of the caustic points (Figure 1.1b), turn out to be quite sensitive to the velocity structure around the discontinuity. For example, the location of cusp-B is sensitive to the velocity gradient in the lower-mantle transition zone right above the 660; the location of cusp-C is sensitive to the velocity below the 660; and the travel-time difference between AB and CD branches is sensitive to velocity jump across the 660.

The triplication method has been applied to constrain the upper mantle discontinuities and core-mantle boundary (CMB) structure [e.g., Lay and Helmberger, 1983; Grand and Helmberger, 1984; Wang and Yoo, 1991; Tajima and Grand, 1995; Tajima and Grand, 1998; Brudzinski and Chen, 2000; Brudzinski and Chen, 2003; Song et al., 2004]. However, due to the sparse distribution of seismic instrumentations, previous studies were largely based on individual or limited seismogram analysis within a large aperture seismological array with which it is difficult to resolve the trade-off between the depth and sharpness of the discontinuity and velocity variation [e.g., Tajima and Grand, 1998; Wang et al., 2006]. To illustrate how a velocity variation around the 660 can influence triplication waveforms, we show synthetic transversals displacements (Figure 1.2a) calculated from three different velocity
Figure 1.1 Diagram of caustic waveforms. (a) Three consecutive phases are expected to arrive at the receiver side. These are the direct AB phase, a wave diving above the discontinuity; the reflected BC phase, a wave reflected from the discontinuity; and the CD phase diving below the discontinuity. (b) Synthetic transversal displacement for an assumed epicentral depth of 520 km. The locations of caustic points are very sensitive to the velocity structure around the discontinuity. (c) Iasp91 model used for the synthetic waveform calculation.

Figure 1.2 (a) Synthetic seismograms at 20° epicentral distance (Δ) for transversal component S wave calculated for iasp91 model and three different models (c). The depth of the event is 520 km. The trade-off between the depth of the interface and velocity variation around the 660-km discontinuity is not well resolved due to the similarity of the waveforms in a single seismogram. (b) Predicted caustic-wave travel-time after aligning seismograms with the epicentral distance. Significant discrepancy appeared in the major features of caustic points, which can be used to discriminate between different models. (c) Different models used for calculation of (a) and (b). Note that the reference model iasp91 (black line) is overlapped with the other three models.
models (Figure 1.2c), which are proposed to characterize deep structure in subduction regions [e.g., Tajima and Grand, 1998; Wang et al., 2006; Wang and Niu, 2010; Ye et al., 2011]. In Model 1, the 660-km discontinuity is depressed to 690 km; in Model 2, the 660-km discontinuity is a layer with a thickness of ~50 km instead of a sharp velocity jump; in Model 3, there is a gentle velocity gradient ($dV/dh = 0.0036 \text{s}^{-1}$) across a layer with a thickness of ~60 km located just above the 660. Over a certain distance, the amplitude of AB and CD phases varies between the models. However, the differences arising from a single seismogram are not clear enough for discriminating the right model (Figure 1.2a). On the other hand, the location of cusp-B, cusp-C, and crossover point O varies significantly (Figure 1.2b). For example, the locations of cusp-B and O are shifted by 0.9° in Model 1 compared with the reference iasp91 model; the AB branch terminates at a much longer distance (~27°) in Model 3.

With the rapidly increasing installation of broadband seismic stations, the record section of seismograms as a function of epicentral distance within a limited azimuth range becomes much more useful and reliable for exploration of the vertical deep-mantle structure [e.g., Wang and Niu, 2010; Ye et al., 2011]. In recent years, more than 800 broadband seismic instrumentations have been installed on mainland China, making it feasible to investigate the deep-earth structure in detail through the dense array technique [Zheng et al., 2010; Niu and Li, 2011].

The velocity structure of the mantle transition zone described in the next section is based mainly on our new findings obtained through caustic waveform modeling [Li et al., 2013]. Here we want to emphasize three important points: (1) In contrast to the traditional single or limited seismogram analysis, we take advantage of a well-constructed fan-shaped profile within a narrow azimuth range (Figure 1.3), and use the major features of the whole stack of seismic records, e.g., the location of caustic points, the crossover distance of the AB and CD branches, and the relative time between the AB and CD phases as well, with the goal of obtaining an

Figure 1.3 Map showing the location of deep earthquakes and regional seismic stations. The beach-ball symbols mark deep earthquake events used for caustic waveform modeling in our paper and previous study [Li et al., 2013], with the red ball indicating the new event. Only events with both P and SH waveforms clearly identified are shown here. The filled triangles represent Chinese regional seismic network stations used for the new event, and the inverted blank triangles represent stations used in our previous study. Brown lines bounded by two sections of black lines highlight portions of the ray traveling below the 660 for event 20130406. The thick blue line AB marks the cross section in Figure 1.8, with “Hinge” indicating the location where the subducting slab encounters the 660 [Li et al., 2008].
accurate seismic image of the upper mantle transition zone beneath northeast Asia. (2) Only events with both P and SH waveforms clearly recorded (Table 1.1) are plotted in Figure 1.3. We used exactly the same station-receiver geometry for both P and SH waves to ensure a close match between the ray paths traversed by the P and S waves. (3) We considered a new dataset associated with one recent deep earthquake (20130406) that occurred near the border of China, Russia, and the Japan Sea (Table 1.1). This dataset, which is characterized by a relatively high signal-to-noise ratio (SNR) for both P and SH waves, was added to verify the overall features of aligned records and the fitness between the observation and theoretical seismograms. We also note that, compared to the dataset of the event 20080519 used in our previous study, the waveform of this new event seems to be a little noisy, and only 35 stations are selected. Therefore, we did not update the velocity models by using all those events simultaneously. The modification to the model is small considering the uncertainty estimates that will be addressed in the following section.

### 1.3. SLAB IMAGE IN THE MANTLE TRANSITION ZONE

The comparison between the observed and synthetic seismograms calculated from our preferred models for the new event is shown in Figure 1.4. A reflectivity synthetic code [Wang, 1999] is applied to generate theoretical seismograms, and a Gaussian wavelet is used to represent the source-time function. The preferred P and SH velocity models obtained from caustic waveform modeling are shown in Figure 1.5.

For the transversal component, the observed AB branch extends to as far as 23°, much farther than ~19° as predicted by iasp91 reference model; the CD branch begins to emerge at a distance of ~12.5°, in contrast to ~10.5° predicted by iasp91. We clearly observe a broadened BOD-zone with significantly delayed AB wave after a distance of 16°, consistent with our previous study. In the aligned vertical waveforms, the AB phase appears to be weaker around a distance of 23°; the CD phase begins to emerge around 13°, much farther than the value of ~11° calculated from the iasp91 model. Due to the limitations imposed by the station coverage, it is difficult to determine exactly the crossover distance of AB and CD phases for both types of waveform. For the detailed description of the caustic waveform modeling and comparison of seismic waveforms of other previous events, we refer the reader to Li et al. [2013].

In Figure 1.4, we can see that locations of caustic points and crossover distance of different branches as well as the relative time difference are matched well over-all for both vertical and transversal components. Although there is some mismatch between the synthetic and observed waveforms, we argue that the local shallow structure and lateral velocity variations may contribute to this discrepancy. Waveform modeling for a 3D velocity structure should be applied to account for lateral velocity variations and for the interaction between the slab and deep mantle.

One of the major features of both P and SH models is that a high-velocity layer with a thickness of ~140 km lies just above the 660-km discontinuity (Figure 1.5), which could not be well constrained from seismic tomographic images [e.g., Huang and Zhao, 2006]. The slab appears to be significantly thickened in the upper mantle transition zone when compared with its initial thickness of ~80–90 km [Zhao et al., 1994] near the Japan trench. To our knowledge, this is the first time the thickness of the stagnant slab in the mantle transition zone is self-consistently resolved from both P and S waveforms. Thickened and broad fast velocity anomalies in the lower mantle beneath subduction zones have been recognized by several authors [e.g., Ribe et al., 2007; Ren et al., 2007; Běhounková and Čižková, 2008]. However, seismic investigation of the thickening of slabs at the bottom of the upper mantle as well as the dynamic implications of this process are quite limited. In the following sections, we will mainly pay attention to this point. We will first present an analysis of the uncertainty for our estimate of the slab thickness; we will then discuss results from numerical simulations in order to explore possible mechanisms and implications of slab thickening in the mantle transition zone.

<table>
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<tr>
<th>Event ID</th>
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<th>Location</th>
<th>Magnitude</th>
<th>Depth (km)</th>
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<td>5.6</td>
<td>513 522 519</td>
</tr>
<tr>
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<td>5.3</td>
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</tr>
<tr>
<td>20130406</td>
<td>06/04/2013 00:29:55.00</td>
<td>42.73 130.97</td>
<td>5.8</td>
<td>563 575 570</td>
</tr>
</tbody>
</table>

*Events are from Li et al. [2013]. Only events with both P and SH waveform used are illustrated here and in Figure 1.3.

†Depth determined by array stacking method [Li et al., 2008].
Figure 1.4 Comparison of the observed (black solid line) and synthetic waveforms (blue dashed line) calculated from the preferred P and S velocity models for the new event 20130406. (a) and (b) refer to vertical and transversal waveforms, respectively.

Figure 1.5 Velocity models obtained from caustic waveform modeling: (a) $V_p$ model; (b) $V_s$ model [Li et al., 2013].
1.4. UNCERTAINTY ESTIMATES OF THICKNESS OF SLAB

Because of the great trade-off between depth of discontinuity and velocity variation, it is hard to provide a quantitative and accurate uncertainty estimate of the obtained velocity model. We adopted here a scheme similar to Brudzinski and Chen [2000]. We explored the model by perturbing the best fitting models until synthetic seismograms showed large misfits in the locations of cusp-B, cusp-C, and O point, or in relative time between AB and CD phases. We focused on two parameters: the thickness of the higher-velocity layer just above the 660 and the velocity variation above the 660. While perturbing one parameter, we tried to keep the other parts of the structure undisturbed. We rejected perturbations that produced misfits larger than ±0.5 s in relative time between AB and CD phases, or 0.5° shift in either cusp-B, cusp-C, or O location relative to the preferred models. We choose these values because significant discrepancies of the major features will appear in the aligned waveforms. However, different cut-off of these values will definitely affect the uncertainty estimation of the slab thickness to some degree.

A series of uncertainty tests was made based on waveform matching. Figure 1.6 shows how the locations of caustic points (cusp-C and cusp-B), crossover distance (point O), and relative time between the AB and CD branches change upon perturbing the thickness of the 660. If we take the above-mentioned rules to be the standard for acceptable models, we arrive at an estimated thickness of the slab to be 140±20 km, which yields a much better constraint to the thickness of the stagnant slab trapped in the mantle transition zone (Figure 1.5).

1.5. DYNAMIC SIMULATION OF SLAB THICKENING

We carried out a series of 2D Cartesian numerical simulations of subduction that can help obtain a better physical understanding of the mechanisms and implications of a thickened slab. The finite-volume-based code YACC was used for the simulation [e.g., Tosi et al., 2013]. YACC has been benchmarked for non-Boussinesq thermal convection [King et al., 2010]. In Figure 1.7, we show three snapshots of the temperature field obtained from a simulation in which a significant thickening of the slab upon interaction with the 660-km discontinuity is observed. These results are from the reference model described by Tosi et al. in Chapter 6 of this volume, to which we refer for a detailed description.

Here we list the essential features of the model. In the framework of the Extended Boussinesq Approximation [e.g., Christensen and Yuen, 1985] with variable thermal expansivity and conductivity [Tosi et al., 2013], we employed a purely thermal convection model in a two-dimensional box with a depth of 2890 km and width of 11,560 km. A 20-km wide weak zone inclined by 30°, with a viscosity of 10^{20} Pa-s is prescribed between an oceanic and an overriding plate. At the trench, the oceanic plate has a thickness of 100 km and a temperature distribution corresponding to that obtained from a half-space cooling model for an age of ~120 Ma. The overriding plate, which is assumed to have a uniform thickness of 100 km and an age of 120 Myr, is kept fixed with a no-slip upper-boundary condition throughout the simulation. After initiating the subduction kinematically by prescribing for the subducting plate a surface velocity of 5 cm/a over a time interval of 4 Myr, we changed its boundary condition to free slip, thereby allowing the slab to fall under its

Figure 1.6 Uncertainty estimation of the slab thickness. We explore the model by perturbing the best fitting models until synthetic seismograms show large misfits in the location of cusp-B (B marks), cusp-C (C marks), and O point (O marks) or in the relative time between AB and CD phases (X marks). If the variation of the upper boundary is taken into account, the uncertainty in the thickness of the slab is estimated to be ±20 km.
own weight. We took into account the effects on buoyancy and latent heat due to the exothermic phase transition from olivine to spinel at 410-km depth, and to the endothermic one from spinel to perovskite at 660-km depth. In addition, we imposed a 10-fold viscosity jump at the 660-km discontinuity and considered pressure-, temperature-, and phase-dependent coefficients of thermal expansion and conduction following the parameterization introduced by Tosi et al. [2013].

Returning now to Figure 1.7, the slab descends along the weak zone and quickly starts to sink nearly vertically because of the choice of keeping the trench at a fixed position. The negative thermal buoyancy drives the slab until it encounters a significant resistance near the 660-km discontinuity (Figure 1.7a), which is caused both by the endothermic ringwoodite-perovskite phase transition and by the imposed viscosity jump. As a result, the slab slows down, and buckling instabilities occur (Figure 1.7b). Upon buckling, more and more slab material keeps piling up above the 660, with the consequence that the overall thickness of the subducted lithosphere is significantly enhanced eventually (Figure 1.7c).

### 1.6. DISCUSSION AND DYNAMIC IMPLICATION

Recent seismic tomography results have revealed a variety of configurations of subducted slabs beneath different subduction regions. Along the western margin of the Pacific plate, the slab beneath the Mariana arc appears to sink into the lower mantle in a nearly vertical fashion [e.g., van der Hilst et al., 1991]; beneath southern Kurile, Japan, and Izu-Bonin arcs, slabs are found to be lying horizontally and stagnate over the 660 [e.g., Fukao et al., 2001; Huang and Zhao, 2006; Li and van der Hilst, 2010; Fukao and Obayashi, 2013]; beneath the Tonga arc, they appear instead to penetrate into the lower mantle after lying horizontally in the mantle transition zone [van der Hilst, 1995]. The spatial and temporal variations in slab strength and the history of subduction forces play an important role in the partitioning of mass across the phase transition boundary, thus leading to the different subduction behaviors [King and Ita, 1995; Billen, 2008].

It is well known that the thickness of the lithosphere does not remain uniform during the subduction process. Slabs tend to thicken while sinking into the lower mantle. This has often been interpreted as evidence for buckling of cold and stiff lithosphere [Gaherty and Hager, 1994; Ribe et al., 2007; Běhouneková and Čížková, 2008; Lee and King, 2011; Tosi et al., Chapter 6 of this volume]. Seismic tomography studies have also revealed the feature of apparent broadening of subducted lithosphere in the middle or lower mantle [e.g., van der Hilst, 1995]. For example, the thickness of slabs may increase from 50 km to 100 km above the 660 to more than 400 km in the middle mantle beneath Java and Central America. Beneath
the Mariana arc and Tonga, the thickness of slabs in the lower mantle is also suggested to increase by a factor of up to five [e.g., van der Hilst, 1995]. Pure compression alone due to the increase of viscosity with depth [e.g., Bunge et al., 1996] cannot account for the apparent observed significant thickening of slabs. A periodic buckling mechanism was proposed to account for the broad fast velocity anomalies beneath the mantle transition zone [Ribe et al., 2007]. The characteristic shapes of the observed anomalies agree well with those predicted by the scaling laws for buckling [Ribe, 2003].

Because of its poor depth resolution, seismic tomography is not able to resolve well the variations in the slab thickness. In the mantle transition zone, the inverted image from first arrivals is even blurred due to complications caused by caustic waveforms. The thickness of initial part of the slab beneath Japan is constrained to be ~80–90 km [Zhao et al., 1994] from travel-time inversion of local, regional, and teleseismic records. This feature is consistent with a slab age of 120 Ma when assuming a potential temperature of 1180°C [Deal and Nolet, 1999]. However, our caustic waveform modeling of both P and S waves revealed a thickened slab trapped in the upper mantle transition zone beneath northeast Asia, which appears to be thickened by 50–60 km at the long-standing segment, indicating ongoing piling or thickening under northeast Asia today.

One of the earliest works on folding of viscous plumes falling vertically onto a density or viscosity interface was that of Griffiths and Turner [1988]. They suggested that buckling might occur when subducted oceanic lithosphere interacts with a density and/or viscosity interface around the 660 in the deep mantle. Our simulations deal with a more realistic case in which the dynamic interface around the 660 in the deep mantle. Our simulation. The compressional stress axes of intermediate-depth and deep earthquakes are all parallel or subparallel to the down-dip motion of the subducting part of slab, indicating a compression-dominated regime in the downgoing slab [Li et al., 2013]. A recent mid-mantle scattered waves study has constrained the viscosity of the upper part of the lower mantle beneath the same region to be in a range of $1.0 \times 10^{22} - 1.6 \times 10^{23}$ Pa s [Li and Yuen, 2014], also suggesting a possible viscosity contrast across the upper-lower mantle boundary.

Second, the accelerated effect of grain-size reduction due to phase transition of ringwoodite in a cold subducting slab [Karato, 2003; Yamazaki et al., 2005] results in a nonlinear diffuse creep deformation. In addition, the grain size in the cold portion of the slab is expected to be less than ~100 μm, compared with an ambient mantle grain size of the order of millimeters. Viscosity of the slab subducting through the transition zone caused by the grain-size reduction alone is estimated to be four to six orders of magnitude lower than that of the surrounding mantle, if the temperature effect is not considered [Yamazaki et al., 2005]. Such a softer slab could easily bend, deform, and pile up above the 660-km discontinuity. Stegman et al. [2010] indeed showed with 3D numerical models of subduction dynamics that only when using a relatively small viscosity contrast of ~100-300 between subducting plate and upper mantle, sinking plates exhibit recumbent folds atop the lower mantle. Furthermore, weak slabs can be maintained by subduction due to the low growth rate of ringwoodite [Yamazaki et al., 2005].

Third, the pressure-, temperature-, and phase-dependent thermal expansivity and conductivity also contribute to the thickening of the slab. Tosi et al., [Chapter 6 of this volume] has demonstrated that the two parameters exert a dramatic impact on the dynamics of subduction. An increased propensity toward locally layered convection is observed, with slabs trapped in the mantle transition zone when both thermal-dynamic parameters are allowed to vary together [Tosi et al., 2013; Tosi et al., Chapter 6 of this volume].

Although not allowed in our simulation, the trench retreat, or the rolling back of the slab might facilitate the quasi-periodic buckling of the slab in the mantle transition zone. Recently, Cizkova and Bina [2013] explored the time-dependent interplay between trench retreat, slab buckling, and slab stagnation. Their 2D simulations show that if the overriding plate can slip freely, an oscillating behavior of the subducting plate can occur, resulting in a subhorizontal distribution of buckle folds above the 660-km discontinuity. Billen [2010] proposed that trench retreat prior to slabs entering the mantle transition zone is a possible mechanism for trapping slabs in the mantle transition zone. We argue that the subducting Pacific slab along the Japan trench can be interpreted as being in a “buckling/folding retreating mode,” as described, for example, by Ribe [2010]. With the continuation of buckling, the overall thickness of the slab above the 660-km discontinuity is enlarged significantly (Figure 1.7c) as inferred from our seismic analysis.
Piling of subducted material could help explain the enigmatic issue that part of the Pacific plate appears to be missing. It has been estimated that in the past 150 Myr, approximately 13,000 km of lithosphere has descended into the Japan and western North American trenches. The total amount of subducted Pacific plate at the Japan trench over the last 50 Myr has been calculated to be ~4000–5000 km in a reference frame fixed with respect to the overriding plate [Engebretson et al., 1992], which is consistent with the assumption of an average subduction rate ~8–10 cm/yr from the early Cenozoic. Tomographic images of the deep mantle presented in a variety of seismic studies indicate that the western edge of the stagnant slab is located approximately 2000–2500 km west of the Japan trench [e.g., Huang and Zhao, 2006; Li et al., 2008; Fukao and Obayashi, 2013], resulting in a severely limited amount of material that may have been subducted (Figure 1.8). Assuming the piling of the slab is uniform along the stagnant segment, the actual trapped length of the slab can be estimated by multiplying the imaged length of ~1300 km by a thickening factor of 1.6–1.8. The buckling and piling of slab in the mantle transition zone can account for at least 800–1000 km of “missing” Pacific plate.

We can also estimate that the age of the west edge of the subducting Pacific slab is not older than 40–60 Myr, even after accounting for the thickening effect of the slab due to buckling. This is consistent with the drifting history of the Pacific plate, which changed its direction from NNW to NW at ~50 Ma [Sun et al., 2007]. In recent years, the destruction of North China Craton (NCC) has been an important topic in the study of continental evolution. Different mechanisms including thermal erosion [e.g., Xu, 2001, 2007; Menzies et al., 2007], delamination [Wu et al., 2002; Gao et al., 2004], Pacific subduction [Niu., 2005], or even the India-Eurasia collision and mantle plume activities have been proposed by different authors [e.g., Liu et al., 2004; Wilde et al., 2003]. Recent studies suggested that the Pacific subduction [Zhu et al., 2012] and the subduction of the ridge between the Pacific and the Izanagi plates [Ling et al., 2013] are the principal triggers for the destruction of the NCC. We argue that the ongoing piling and stagnation occurred after 50 Ma, and thus the stagnation of the Pacific slab observed so far has no direct relationship to the destruction of the NCC, which took place with a peak age of ~120–125 Ma [e.g., Zhu et al., 2011; Ling et al., 2013]. At least it cannot be the direct cause of the reactivation of the stable North China Craton.

The continuous piling up of slabs may have important implications for the fate of the descending lithosphere under northeast Asia (Figure 1.9). Tomographic images are only a snapshot in time and that slab morphology and mass flux into the lower mantle may be a transient phenomenon [Griffiths et al., 1995]. With the ongoing piling of the trapped cold lithosphere, we argue that the slab might be on the margin of instability. The 3D
numerical simulations of mantle convection [Honda et al., 1993b] indicate that, when considering the olivine family of phase transitions, cold material is likely to accumulate at the phase-change boundary between the upper and lower mantle, eventually resulting in a large-scale gravitational instability. To the west of the Japan trench, convection seems to be temporarily hindered by the phase changes. A large volume of cold and dense material has been accumulating at the 660-km discontinuity. This could in turn rapidly sink into the lower mantle, leading to a mantle avalanche. The occurrence of instability, which may be in the future 5–10 Myr, would be accompanied by a flush of volcanic activity [Honda et al., 1993a; Honda et al., 1993b], precipitated by mantle upwellings emanating from the lower mantle (Figure 1.9).

1.7. CONCLUSIONS

The motivation of this paper is to stimulate more multidisciplinary work revolving around the study of the structure and dynamics of the mantle beneath northeast Asia, where significant material appears to be stagnating in the transition zone. We have partly addressed the issue employing caustic waveform modeling to image the slab. We inferred the presence of a long stagnant slab with a thickness around 140 km, possibly indicating ongoing piling of subducted oceanic lithosphere. In addition, we showed numerical simulations indicating that the subducting lithosphere exhibits buckling instabilities due to a combination of viscosity jump across the upper-lower mantle interface, phase transitions, and variable thermodynamic parameters that naturally lead to a significant thickening of the slab. This thickened slab in the mantle transition zone can partly account for the missing volume of the Pacific slab subducted beneath the Japan trench for the last 50 Ma. We also speculate that such a thickened stagnant slab might be close to an impending instability with a rapid flush of cold material possibly occurring in the next few million years.

ACKNOWLEDGMENT

This work is supported by NSFC (J. Li, Grants 41322026 and 41274065). We thank Gabriele Morra, Taras Gerya, Weidong Sun, and Yongtai Yang for helpful discussion and Scott King for comments and suggestions that helped improve the manuscript significantly. An anonymous reviewer provided constructive and critical suggestions. Center of China National Seismic Network at Institute of Geophysics, China Earthquake Administration provided the seismic data. GMT software was used in figure plotting. This research was also supported by CMG and Petrology program of US NSF. N. Tosi acknowledges support from the Helmholtz Gemeinschaft (project VH-NG-1017).
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